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Observation of local cloud and moisture feedbacks associated with high ocean and desert surface temperatures

Moustafa T. Chahine

Jet Propulsion Laboratory, California Institute of Technology, pasadena, California, 91109, USA

New data on clouds and moisture, made possible by re-analysis of weather satellite observations, show that the atmosphere reacts to warm ocean pools in the Western Pacific Ocean with increased moisture and cloudiness, suggesting a negative feedback limiting the rise in sea-surface temperature. The reverse was observed over hot deserts where both moisture and cloudiness decrease, suggesting a positive feedback perpetuating desert conditions. These observations reveal complex dynamic-radiative interactions which show why this problem has been debated for over twenty years.

Prediction of greenhouse warming depends on knowledge of how clouds and atmospheric moisture interact with the surface and the general circulation of the atmosphere, a subject which has been widely debated[1,2,3]. Not only is there no simple theory to explain the interaction with precision, but we do not yet have the basic observational evidence required to validate such theories. Clouds and moisture show strong internal feedbacks and at the same time they amplify and dampen the global and local energy balances. Because clouds reflect solar energy, a simple argument could be made that as the surface temperature of the earth increases, evaporation from the oceans increases, cloud formation is, in turn, intensified and the net energy deposited at the Earth's surface decreases. By this mechanism clouds provide a negative feedback. Clouds also intercept long wavelength radiation from the surface, retaining it in the lower atmosphere, a positive feedback. The sign and amplitude of changes in the atmospheric and surface energy balances and the concurrent effect of atmospheric moisture on greenhouse warming are difficult to explain by radiation processes alone. Dynamical processes (Fig. 1) in the atmosphere exert controlling effects, as well. Accurate observations are needed, models must be refined and validated, and diverse mechanisms must be invoked to explain these phenomena.

Since clouds and moisture vary rapidly as a function of space and time, it is necessary to observe them simultaneously in order to ascertain their links to each other and to the environment with which they interact. Although this has been difficult in the past, the data set developed for this study used coincident observations from two instruments on the same satellite. The resulting global temperature, moisture and cloud data simultaneously characterize the airmass and the underlying surface in each field of view. These characteristic parameters were derived daily during the months of January, April, July and October, 1979.

From this new data we have observed that atmospheric moisture invariably increases at all atmospheric levels as a function of increased sea surface temperature over tropical oceans. This increase in moisture above the boundary layer is most critical in determining the net effect of water vapor on greenhouse warming {A. Arking, manuscript submitted}. Specifically, we have observed strong interactive processes between moisture and clouds over warm pools in the western Pacific. These interactive processes persist over long periods of time under various global and regional. circulation patterns and appear to regulate the local environment through radiative-convective mechanisms. However, over hot and dry desert areas, both moisture and cloudiness decrease as a function of increased surface temperature (above 304 K) providing a positive feedback mechanism which tends to maintain the desert conditional. Intriguingly, the observations show a common critical surface temperature for both oceans and land; the distribution of atmospheric moisture is observed to reach a maximum value when the surface temperature approaches 304 ±1K, then decreases beyond that temperature.

This paper presents new observational data and discusses the implications for climate studies with emphasis on two specific regions: warm ocean pools in the western Pacific and hot desert spots in Australia. The goal is to provide basic new information needed to improve our understanding of the processes which sustain the local surface temperatures of ocean and desert regions. Climate models with diverse approaches to parameterizing clouds and moist convection must be able reproduce these observations as a test of their accuracy and ability to make reliable predictions.

Observations

The data derived for this study come from the High Resolution Infrared Sounder (HIRS) and the Microwave Sounding Unit (MSU) flown on the NOAA low-Earth-orbiting operational weather satellites since December 1978[5]. NOAA has derived global atmospheric temperature and moisture data since 1979; in the early years, NOAA analyzed the HIRS/MSU observations using a statistical approach[6] which employed empirical relationships between satellite observations and atmospheric parameters gathered from radiosonde and rocketsonde reports. However, the results of this approach do not meet the accuracy and consistency requirements of this study. To satisfy the study requirements, a completely analytical method[7] was applied to re-analyze the HIRS/MSU observations, starting with the year 1979[8]. The infrared and microwave satellite weather data permit both the "clear-sky" and the "cloudy-sky" properties to be derived at all times[9] even though the atmosphere is never free from clouds and haze (see Box 1) .

Specifically, the derived parameters consist of atmospheric profiles of temperature and precipitable water vapor (PWV), sea-surface temperature (SST), land-surface temperature (LST), the effective cloud infrared opacity (α) and the cloud-top height and temperature. PWV is simply the mass of water vapor in a column of air bounded by two pressure levels. The effective infrared cloud opacity in a field of view is defined as the fraction of energy (between 8 and 15μm) intercepted by the clouds, that is emitted by the surface and atmosphere below the clouds. We define, also, the clear-sky radiance (F_{cl}) as the radiant energy emitted by the surface and the atmosphere, that is transmitted through the clouds and through openings in the clouds. If we define the cloud radiance (C_{cd}) as the energy emitted by the clouds and the atmosphere above them for full overcast conditions, we can then relate F_{cl} and C_{cd} to the observed radiance (F_{cd}), i.e., the cloudy-sky radiance in a given field of view, as[9]

$$F_{cd} = (1-\alpha)\, F_{cl} + \alpha\, C_{cd}$$

The data for January 1979 were specifically selected for the illustrations in this paper because their accuracy had already been determined by careful comparison with *in situ* observations under diverse cloud and surface conditions (see Box 1). An important limitation of this data set is that it does not contain the cloud properties in the visible part of the spectrum, because the visible channel on the HIRS instrument was not calibrated. This makes it impossible to determine the net radiative effects of the clouds.

Warm Pacific ocean pools

The sea surface temperature (SST) values derived from the space observations are the radiating temperature of the ocean "skin surface" and not the "bulk" SST values usually reported by ships. The derived SSTS showed several clusters of warm water pools in the tropical Pacific between 20"N/20°S latitude and 150°E/180° longitude. A representative cluster to the northeast of Australia is shown in Fig. 2a for daytime observations on January 4, 1979. These warm pools were observed every day in January 1979. They stretched across hundreds of kilometers and meandered slowly in the western Pacific, fluctuating slightly in size and intensity. A strong correlation was noted between changes in the precipitable water vapor (PWV) and the infrared cloud opacity (α) with the temperature gradient across the warm pools. This correlation became weaker on certain days, but never reversed its sign. Thus, in this region of deep convection in the tropical western Pacific, the warm pools and the airmass above cannot be considered in isolation from each other.

The variations of the PWV as a function daytime SST across the warm pools and the sea surrounding it are shown in Fig.3 for the month of January 1979. Since the accuracy of the derived SST is estimated to be ±0.5°C, a moving average of the corresponding values of PWV was calculated across an interval of 0.5°C. This step reduced the noise (or scatter) of the individual daily observations of PWV by about 20%. The resulting "smoother" distributions correspond to variations of PWV in four layers bounded by 100, 300, 500, 700 mb and the surface. The total PWV is given in Fig. 2e. These results clearly show a very strong coupling between increases in moisture at all atmospheric levels and the increase in SST across the warm pools. Of distinct interest, the PWV at the higher altitudes (in Fig.3a) increased faster than in any other atmospheric layer. For example, as the SST across the warm pools increased from 300 K to 305 K, the PWV between 700 mb and the surface increased by 20%, while the corresponding increase between 100 and 300 mb was 100%. At SST values above 305 K, the distribution of the total PWV always decreased (Fig. 3e). However, the small number of observations above 305 K (as shown in Fig. 3f) should be noted. Fig. 3e shows also the distribution of the corresponding effective infrared cloud opacity, α, and its relationship to the variation of moisture with SST. In the range between 300 and 305 K, the effective cloud opacity increased by 40% (as shown in Table 1). The use of a moving average of α reduced the scatter of individual daily observation by ±50%.

These high and opaque clouds and the associated PWV were observed every day of the study. Their distributions depended not only on the highest value of the SST but also on the spatial gradient across the warm pool. The controversy surrounding this problem has been debated extensively[3,10-14]. Lindzen[10] proposed that the flow field across the warm pool provides convergence of warm humid air in the boundary layer and the excess energy of condensation heating in the clouds is removed by the divergence of the flow of air at the higher levels. However, in spite of the flow divergence at the upper levels, the moisture in the upper troposphere was observed to increase with increasing SST[11,12]. Furthermore, Ramanathan and Collins[3] proposed that the sharp increase in high level clouds over 'the warm ocean pools acts both connectively and radiatively to produce a negative feedback which contributes to limiting the SST to 305 K. Other dynamical processes both in the ocean and the atmosphere have been proposed to account for the observed local SST limit[13,14].

All these feedbacks were observed daily during the four months of the study; they appear to be local (or regional but not global) and are controlled by both the spatial gradient of SST and radiative-dynamic atmospheric processes. For example, warm pools in the Indian Ocean (shown in Fig. 2a) which exhibited lower SST gradients showed smaller increases in both PWV and α from the edge to the center of the pool. The main common feature was at SST values of 305 K, where both α and PWV were always observed to undergo a sudden and sharp increase from their levels at 303 K.

The observational data also indicate that the distribution of atmospheric water vapor remains correlated with the surface temperature everywhere in the tropical oceans[12]. To validate this, the distribution of PWV was examined for SST values between 299 and 300 K in the western Pacific (150°E to 180° and 20°N to 20°S), eastern Pacific (180° to 150°W and 20"N to 20°S) and the Indian Ocean (60°E to 90"E and 0° to 20°S). This value of SST was selected for comparison because it sits at the outer edge of the warm pools and, moreover, it is among the most frequently observed value of SST in each of the three regions. In all three regions the total PWV was equal to 3.9 ±0.2 g/cm^2 and the PWV in the 100 to 300 mb layer was 0.015 ±0.001 g/cm^2, well within the observational uncertainties. In addition, the rate of increase of PW at higher altitudes was invariably higher than that near the surface in all three regions.

Hot desert spots

The data shown in Fig. 4 describe the distribution of moisture and clouds as a function of daytime land surface temperature (LST) of hot desert spots. Because the error in the derived land surface temperature is estimated to be at least ±1°C, we applied a one degree moving average to the individual data points. As in the case of oceans, this step reduced the scatter in the individual daily data points by about 20% for humidity and 50% for the cloud infrared opacity. As the surface temperature increased above 303 **K** the total PWV (Fig. 4e) decreased very rapidly accompanied by a sharp decrease in cloudiness. However, the decrease in PWV occurred mostly near the surface in the boundary layer. Between 300 and 320 K the PW decreased by 30% in the layer between the surface and 700 mb and 20% between 300 and 100 mb. At the high desert temperature of 320 K, the effective cloud opacity decreased to a minimum of 0.19 in contrast to the maximum value of 0.39 observed at the center of ocean warm pools.

The surface albedo, vegetation, soil dryness, and the general circulation of the atmosphere all interact to affect the local cloud cover, moisture and rainfall over land. At this time, there is no scientific consensus on the causes and feedback mechanisms that trigger and maintain desertification. However, the feedback model proposed by Charney[1] remains the leading explanation of the links between desert surface conditions and atmospheric moisture, clouds and rainfall. According to the Charney model, the high surface albedo of dry, light and arid deserts decreases the net radiation at the surface and increases the radiative cooling of the air above, as observed in Table 1. As a result, air sinks in order to maintain thermal equilibrium, and cloudiness and precipitation are diminished. Low level subsidence reduces moisture in the boundary layer and thus limits the moisture available in the atmospheric column, as observed in Fig.4. Furthermore, reduced moisture and precipitation adversely affect vegetation which in turn increases surface albedo further. This positive feedback was confirmed in subsequent modelling studies[15-18] which showed that surface aridity and the accompanied subsidence and low atmospheric humidity contribute to the persistence of desert conditions.

The results in Fig. 4 were observed, with noticeable regional variations, over hot tropical and subtropical deserts such as the Sahara, Kalahari and central Australia, higher latitude deserts in continental interiors such as the Gobi and Kazakhstan in central Asia, as well as deserts found on the west coasts of Africa and south America such as the Namib and the Atacama. However, unlike the general (semi-permanent) nature of the observations noted over the western Pacific warm pools, the results obtained over the hot desert spots showed dependence on seasons. This is to be expected because most desert regions are located near the descending branch of the Hadley cells which vary with seasons and sometimes become weak and even vanish. In addition, seasonal changes in insolation reduce the maximum LST of desert hot spots to less than 310 K.

Longwave fluxes

A one-dimensional low resolution radiation model[19], LOWTRAN 7, was used to examine the relationship between longwave fluxes and surface temperature on the basis of the temperature, moisture and cloud data derived for this study. The cloud infrared opacity, α, was assumed constant across the spectrum. The specific humidity profile (in g/kg) was determined under the assumption that the relative humidity is constant in each of the four observed layers and a fixed value of 0.002 g/kg was assumed above 100 mb. Table 1 gives a representative set of the data used for the ocean and desert flux computations.

The contrast in the cloud-moisture relationships over ocean and desert regions, illustrated in Fig. 5, is very striking. Over the ocean, the outgoing longwave radiation (OLR) for clear-sky conditions increased with increasing SST from 290 to 300 K, levelling off at 300 K in spite of the increased surface emission (Fig. 5a). This is a result of the sharp rise in atmospheric moisture above 300 K, resulting in enhanced atmospheric absorption at the expense of OLR. By contrast, hot desert spots (Fig. 5b) show a continuous increase in OLR which is directly traceable to the sharp increase in the desert surface temperature (the temperature lapse-rate effect) and the simultaneous decrease in water vapor absorption.

The atmospheric greenhouse warming increased with surface temperature for both ocean and land, however for different reasons. The results for clear-sky conditions (Fig. 5c) compare and contrast the atmospheric greenhouse warming due to increased atmospheric moisture over the ocean with that linked to reduced moisture over desert surfaces. Over deserts, while the fraction of the surface emission absorbed by the atmosphere decreased because of reduced moisture, the net emission from the surface increased much faster (as σT^4), resulting in a net increase in the atmospheric greenhouse warming. Nevertheless, it is very interesting to note that for clear-sky conditions, the atmospheric greenhouse warming is nearly identical over both land and oceans for the same surface temperatures between 290 and 305 K. The infrared cloud effect, shown in Fig. 5d, amplified the atmospheric greenhouse warming even further, especially over the ocean.

Effect on climate modelling

Although the use of observations to study and understand climate feedback processes is indispensable, it is generally impossible to separate causes and effects without the application of models. At issue for both oceans and deserts is the need to isolate and quantify the relative contributions of surface temperatures, atmospheric moisture and clouds to the feedback processes.

Atmospheric general circulation models are very sensitive to changes in the amount of atmospheric water and its vertical distribution. The observations for ocean warm pools clearly show that water vapor increased at all levels throughout the troposphere when local warm ocean pools formed, thereby considerably increasing the infrared opacity of the atmosphere and enhancing the infrared atmospheric warming. The fact that this relationship between clouds, moisture and SST was observed on a daily basis for four months in four seasons clearly suggests that the local convective circulation over these warm pools persists under various conditions of the general circulation of the atmosphere. Yet, general circulation models must account for it because of its impact on climate. Before models can be used they must first demonstrate sufficient realism by reproducing the observations presented here.

General circulation models must also reproduce the local desert observations. Above the desert surface the mean atmospheric temperature remained essentially unchanged (Table 1) while the desert surface temperature approached 320 K. Under these conditions the lapse rate of temperature near the surface will likely exceed a critical value depending on temperature, moisture and pressure. When this happens the atmosphere becomes unstable and convective adjustment to reestablish atmospheric stability becomes very dominant. In the subtropics this adjustment act to reinforce the descending branch of the Hadley circulation and significantly intensify the positive feedback conditions over the deserts.

The results presented here are local or at best regional. How the observations are linked together on a global scale is not yet well understood. Determination of these links will require interactive coupling between modelling and observational data. Additional data will be required on the optical properties of the clouds. Until these requirements are met, the controversy over the cloud-moisture feedbacks will likely continue. Certainly, nature is not so simple as to be explained by infrared observations alone.

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FIGURE CAPTION LISTING

Fig.1 Computer generated three-dimensional cloud image illustrating the topography of cloud-tops over the Equatorial Pacific ocean and Australia for the month of January 1979. The cloud data were obtained from infrared and microwave observations measured by NOAA weather satellites. Cloud-top height is denoted by color with red for high, green for middle and blue for low level clouds. The strongest color shades represent the most opaque clouds. Note the sparse clouds over the bulk of the Australian desert. The vertical scale has been stretched by a factor of 120 to enhance the clouds vertical structure. The three-dimensional image displays land contours projected on the clouds as if the observer is looking from the south. This cloud formation represents an important convective weather system associated with warm ocean pools reaching a temperature of 305 K. The lower image was generated by T. Van Sant and the Geosphere Project using data from the NOAA Advanced Very High Resolution Radiometer.

Suggested caption for this figure, if used as a cover image:

3-dimensional data visualization showing the long ridge of high clouds stretching across the Western Pacific and the cloudless Australian desert in January 1979. Red, green and blue represent high, middle and 1 Ow level clouds . These high clouds illustrate an important convective weather system connected with warm ocean waters . The lower image is provided by T. Van Sant and the Geosphere Project.

Title options: **Cloud climate connections**
Cloud climate control

Fig. 2 Location of the observed warm ocean pools in the equatorial Pacific (a) and the hot desert spots in Africa and Australia (b) as observed on January 4, 1979. Linear kriging24 was applied to the observational data to underscore the patterns of spatial continuity and enhance the visual appearance of the clusters of warm and hot points.

Fig. 3 Daytime dependence of precipitable water vapor (PWV) in four atmospheric layers (a, b, c, and d) on variations of sea-surface temperature (SST) across the warm pools and their surroundings (SST <300K) in the equatorial Pacific ocean during the month of January 1979. The total PWV and the corresponding distribution of the effective infrared cloud opacity, et, are shown in (e) . A moving average of 0.5°C was applied to both PWV and α to reduce uncertainties in the accuracy of SSTS. The number of observations corresponding to each value of SST is given in (f).

Fig. 4 Daytime variations of precipitable water vapor (PWV) in four atmospheric layers (a, b, c, and d) as a function of land-surface temperature (LST) across the hot spots and their surroundings (LST <303 K) during the month of January *1979.* The total PWV and the corresponding distribution of the effective infrared cloud opacity, α, are shown in (e) . A moving average of 1°C was applied to both PWV and α to reduce uncertainties in the accuracy of LSTS. The number of observations corresponding to each value of LST is given in (f) . The data are truncated at *320* K because of large uncertainties in the accuracy of very high LST values and the small number of corresponding observations.

Fig. 5 Calculated daytime outgoing longwave radiation (OLR) at the top of the atmosphere for the cases of the warm ocean pools and the hot desert spots discussed in Figs. 3 and 4. The clear-sky OLR, the (observed) cloudy-sky OLR and their difference, the cloud forcing, are shown in (a) and (b) . The atmospheric greenhouse warming, defined as the difference between the surface emission σT^4 and OLR, is shown in (c) for clear-sky conditions, and in (d) for cloudy-sky conditions.

TABLE CAPTION LISTING

Table 1. January 1979 Mean monthly atmospheric parameters for a representative set of ocean and desert surface temperatures. The mean monthly values are the average of all the daytime observations collected within ±0.5°C around the indicated surface temperature.

BOX 1 Accuracy and characteristics of the observational data

The Earth's atmospheric temperature and humidity profiles have been observed by infrared and microwave satellite sounders for more than two decades. The HIRS/MSU instruments are multi-channel devices which measure the Earth's outgoing radiance at frequencies selected to characterize different levels of the atmosphere. The HIRS instrument has *19* infrared channels between 3μm and 15μm and one (uncalibrated) channel in the visible. The observed radiances in these infrared channels are affected by a number of atmospheric and surface parameters such as temperature, moisture, clouds and ocean and land surfaces. The MSU has 4 channels in the 50 GHz region which are influenced mostly by atmospheric temperature and the surface. Unlike the infrared channels, the microwave channels are not influenced by most types of clouds.

The algorithm employed here is based on the relaxation method of solution of the full radiative transfer equation[7]. By combining the analysis of the microwave and infrared channels, the algorithm produces accurate temperature and moisture profiles even in the presence of clouds, without requiring any field of view to be clear[20]. Basically, the algorithm is iterative and 'involves finding atmospheric and surface parameters which, when substituted into the radiative transfer equation, simultaneously satisfy the infrared and microwave observations, to within a specified level of convergence accuracy. All solutions that meet the convergence criteria are accepted; the rest are rejected. In general, solutions are accepted up to an effective cloud opacity in the field of view of 80%.

The accuracy of the derived temperature profiles with respect to radiosondes varied from 1.8°C for the clearest cases to 2.0°C for the cloudiest conditions (Susskind, J., M.T. Chahine, P. Piraino and L. Iredell, manuscript submitted). The mean monthly distribution of the PWV shows agreement near 20% with co-located radiosondes[21], with the largest discrepancies occurring over land.

The accuracy of the mean monthly SST is estimated to be ±0.5°C from a study[22] conducted for a special observing period in 1982. Validation studies of LST, the effective cloud opacity and cloud-top heights are more difficult to achieve directly because of lack of field measurements. The error in LST is estimated to vary from ±1°C for low surface temperatures to *3°C for temperatures near 320 K. Above 325 K the accuracy of LST is unknown. It is very important to point out that the space derived SST and LST refer to the radiation temperature of the surface, which could differ significantly from the customary *in situ* measurements. As to the effective cloud opacity α, indirect validations[23] obtained by comparing the computed outgoing longwave radiation (OLR) for the retrieved HIRS/MSU results with that from the Earth Radiation Budget data for 1979 showed a spatial standard deviation of approximately 6 W/m^2. This agreement is good evidence that the retrieved cloud and surface parameters are sufficiently accurate because of the large dependence of OLR on surface and cloud parameters. For example, for a scene with high clouds and *0.50* cloud opacity, an error of 6 W/m^2 can result from an error of 0.5 km in cloud-top height and only 0.02 in cloud opacity, assuming that these are the only sources of error.